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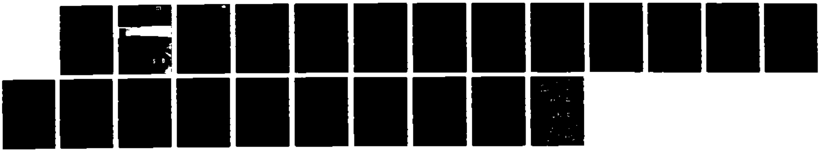

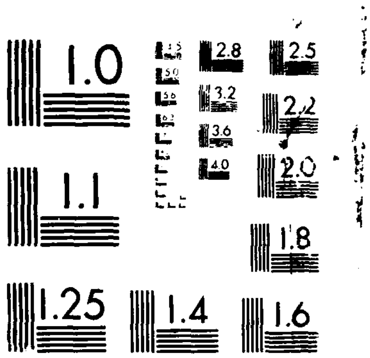

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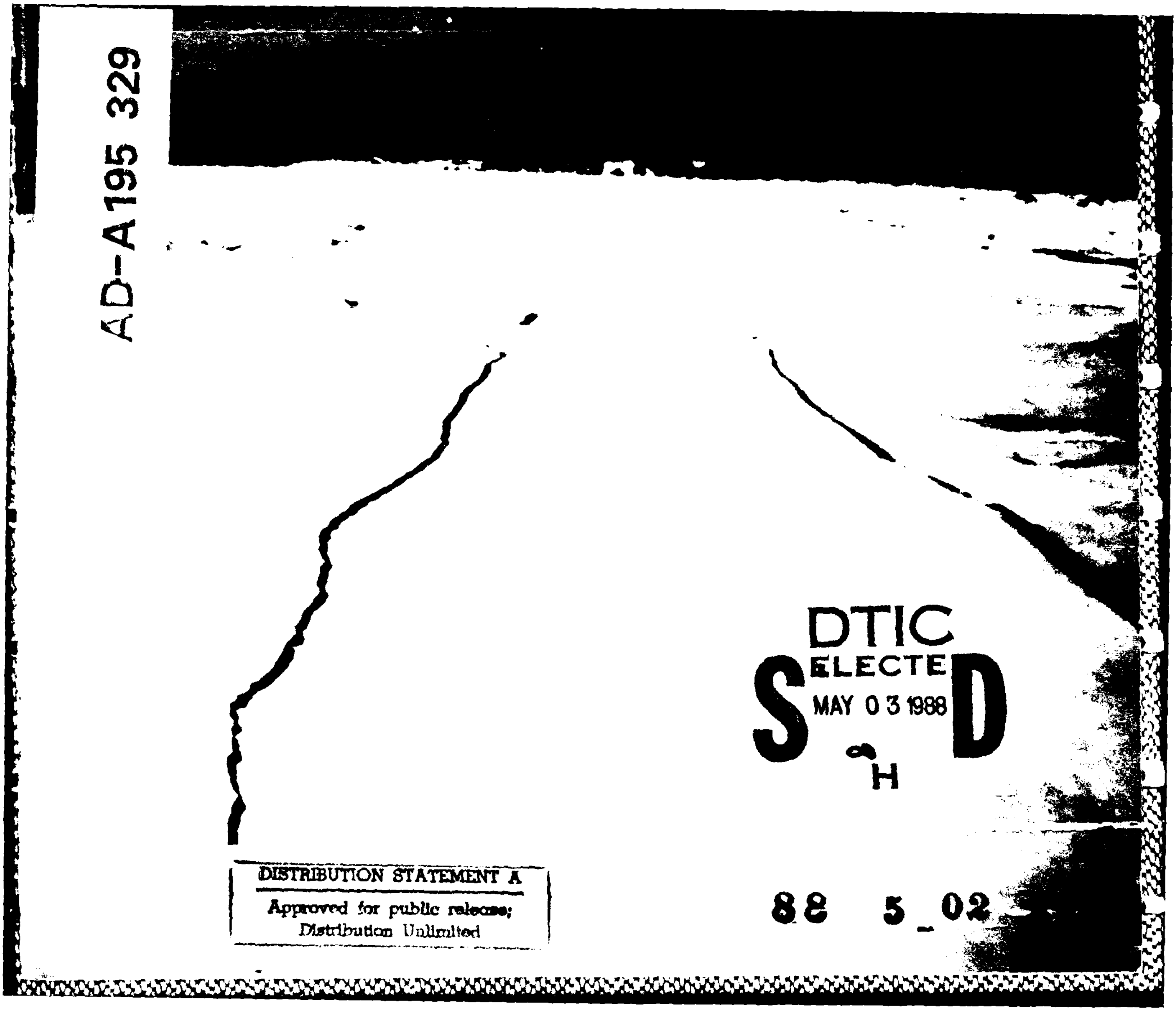

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Options for management of dynamic ice breakup on the Connecticut River near Windsor, Vermont

Michael G. Ferrick, George E. Lemieux,
Patricia B. Weyrick and Warren Demont

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The Cornish-Windsor bridge is the longest covered bridge in the United States and has significant historical value. At a large peak flow, dynamic ice breakup of the Connecticut River can threaten the bridge and cause flood damage in the town of Windsor, Vermont. Throughout the 1985-86 winter we regularly monitored ice conditions, including a midwinter dynamic ice breakup on 27 January. We conducted controlled release tests over the operating range of the turbines at Wilder Dam upstream during both open water and ice cover conditions. These data and observations were analyzed in light of more than 60 years of temperature and discharge records. Our analysis indicates that river regulation presents alternatives for ice management that would minimize the probability of bridge damage and flooding during breakup. The flow can be regulated early in the winter to promote the growth of a stable ice cover, minimizing the total ice production in the reach. In the weeks prior to breakup, sustained releases and above-freezing air temperatures cause melting, weakening and gradual breakup of the ice, greatly reducing the flooding potential. Also, it is possible to produce a controlled ice breakup prior to an imminent natural event at lower stage and discharge. All of these ice control alternatives have associated power production costs.

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PREFACE

This report was prepared by Michael G. Ferrick, Hydrologist, George E. Lemieux, Physical Scientist, Patricia B. Weyrick, Physical Science Technician and Warren Demont, Physical Scientist, of the Snow and Ice Branch, Research Division, U.S. Army Cold Regions Research and Engineering Laboratory. Funding for this research was provided by DA Project 4A762730AT42, *Design, Construction and Operations Technology for Cold Regions*, Task CS, Work Unit 003, *Winter Battlefield River Mechanics*, and by the ILIR (In-House Laboratory Independent Research) program at CRREL entitled "Analysis of River Ice Breakup." Warren Demont performed this work on a sabbatical leave, and his support was provided by the Dresden School District.

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Options for Management of Dynamic Ice Breakup on the Connecticut River Near Windsor, Vermont

M.G. FERRICK, G.E. LEMIEUX, P.B. WEYRICK, W. DEMONT

INTRODUCTION

Initially constructed in 1796, the Cornish-Windsor covered bridge was destroyed by the Connecticut River in the spring of 1824, in 1849, and again on 3–4 March 1866 (Childs 1960). The loss of the third bridge in 1866 was specifically attributed to ice breakup. The present structure was constructed in 1866 at a higher elevation above the river than the previous bridges. Rawson (1963) reports that ice jam floods damaged this bridge in the spring of 1925, 1929, 1936 and 1938. Significant damage occurred again on 14 March 1977 from ice impacts on the upstream side, up to a meter above the bottom of the bridge. The water levels associated with ice damage to the bridge also caused significant flood damage in the town of Windsor, Vermont.

The highest water levels on the Connecticut River near Windsor typically occur during a dynamic ice breakup. Dynamic ice breakup is the rapid failure of an ice cover that occurs during periods of intense runoff when the river is covered by intact ice. There are no reports of damage or threatened loss of the bridge at the present elevation resulting from open water floods with much higher peak discharges.

In this report we characterize dynamic ice breakup on the Connecticut River near Windsor, identify the combination of conditions that produce extreme breakup flood events, and provide data that are preliminary to quantitative predictions of ice breakup behavior. These studies suggest methods of ice control that use flow control at the existing dams on the river to alleviate the threats to the bridge and to the town. The essence of these methods is to minimize ice production over the winter, to melt and weaken the ice prior to breakup, and to produce a controlled ice breakup at the minimum possible water levels prior to an uncontrolled natural event. Additional studies and river monitoring are needed to optimize the river control operations, yielding maximum ice control with minimum loss of power production.

BACKGROUND

The energy gradient of a river is a dimensionless parameter that quantifies the rate of energy dissipation in the flow. The water surface gradient is generally a good indicator of the energy gradient. Changes in the flow release at a dam create long-period river waves (Ferrick 1985). Both the water surface and energy gradients on the front of a river wave can be significantly larger than those found during steady flow conditions at a comparable depth or discharge. The energy gradient varies with stage, discharge, and ice conditions, and can be greatly affected by breakup. The energy in the flow of a river is coupled into an ice cover by a number of mechanisms. Dynamic ice breakup occurs when the hydrodynamic forces on the cover, related to the energy gradient of the river, exceed the resistance provided by the ice strength and points of support (Ferrick et al. 1986b).

Thick and competent ice covers with strong "hinge" connections at the banks have great resistance, and require high energy gradients to initiate breakup. High energy breakup conditions are an extreme that produce the highest water levels. In a high energy dynamic breakup the downstream movement of a single breaking front, generally associated with the front of a river wave, is common. While moving, the breaking front progresses rapidly, but it may stall for a period of time, forming an ice jam. Jams produce water levels that are even higher than those of a moving breakup, and a sudden jam release can be destructive. Locations where the

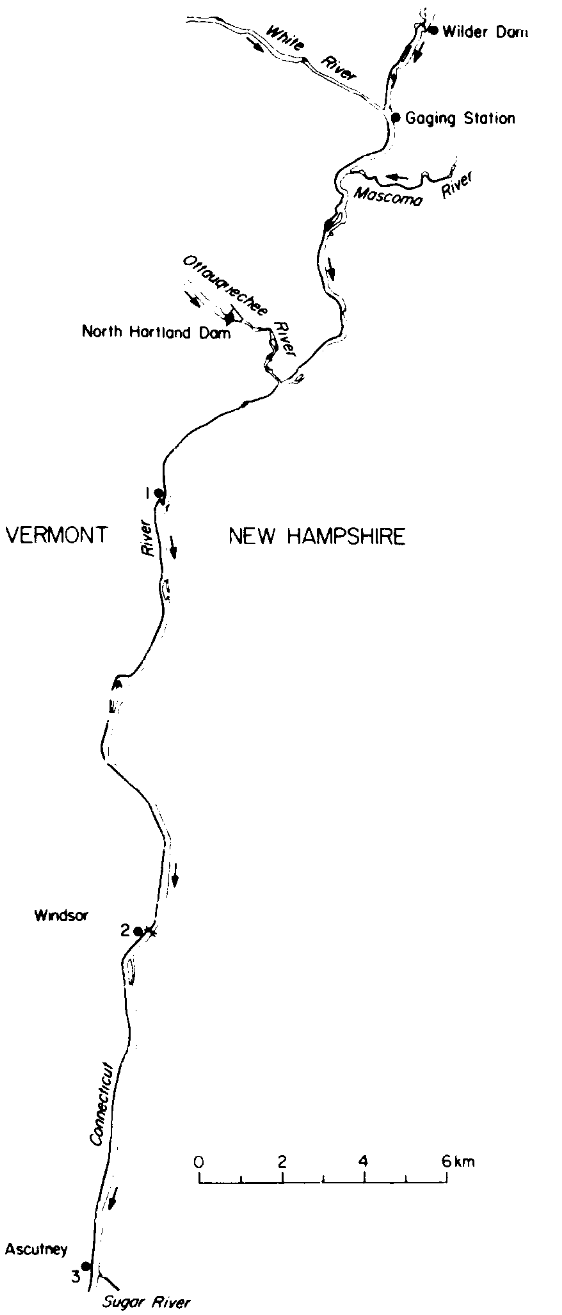

Figure 1. Connecticut River study reach including the major tributaries. Station 2 is located at the Cornish–Windsor covered bridge.

energy gradient diminishes, such as at the upstream end of impoundments of the river, are favorable for ice jam formation.

The flow of the Connecticut River in the Windsor area is controlled by Wilder Dam upstream and Bellows Falls Dam downstream. Our study has focused on the 35-km reach from Wilder Dam downstream (Fig. 1). The river is free-flowing over most of this reach before slowing as it enters Bellows Falls reservoir. Wilder Dam has a drainage basin of 8780 km^2. The White River, with a basin size of 1820 km^2, is the primary tributary in the study reach, entering about 2 km downstream of the dam. At an average discharge of 200 m^3/s the Connecticut River in the study reach varies between 100 and 200 m in width and has a mean depth range between 1.5 and 3.0 m. As Wilder Dam does not generally pass large quantities of ice, the uncontrolled White River is the only significant ice source at breakup external to the reach itself.

We monitored the ice conditions regularly throughout the 1985–86 winter, including a midwinter dynamic ice breakup on 27 January 1986, and conducted a series of controlled release tests over the operating range of the turbines at Wilder. We established temporary data collection stations for the test period, at regular intervals on the river, to supplement the ongoing data collection at the gaging station and the dams (Fig. 1). Station 1 is located immediately upstream of Sumner Falls, a natural river control feature. Station 2 is located at the Cornish–Windsor Bridge, and station 3 is located several kilometers into the pool of Bellows Falls Dam. In addition, White River flow data are collected at a permanent gaging station at White River kilometer 11.9 (WRK 11.9).

The controlled release tests provided data that describe the response of the river to a range of unsteady flow inputs, and the effect of an ice cover on this response. Additionally, these data can be used to calibrate and verify a numerical hydraulic model of the river (Ferrick et al. 1986a and 1986b), which in turn is necessary for design of a controlled ice breakup study or to analyze a natural breakup. The 1986 midwinter breakup occurred with an intact ice cover on the river. Together with the numerical model, these data and observations provide a case study that indicates the approximate energy gradient needed to cause breakup of a well-developed and competent ice cover on the Connecticut River.

The river, ice and breakup observations from the winter of 1985–86 are most valuable for assessing the flooding/bridge damage problem when they are considered as part of a long period of records. Several parameters that affect the breakup water levels can be obtained from the historical records, including the peak breakup discharge, the accumulated subfreezing winter temperatures, the relative timing of the White River breakup, the occurrence of a prior midwinter ice breakup, and the suddenness of the onset of high discharge and warm air temperatures. The contributions of each of these parameters to the largest historical events are evaluated in this report by using air temperature and river stage–discharge records. Knowledge of the conditions that

Table 1. Historical data for years of high breakup flow on the Connecticut River. The data are grouped according to whether or not Cornish–Windsor bridge damage was reported.

Year	Peak flow date of breakup event	Peak daily avg. discharge (ft³/s)	Peak daily avg. discharge (m³/s)	Discharge rank	Hydrothermal melting (m³/s-days)	Freezing °C days	Cold rank	Melting °C-days through peak Q	Precip. in warm period (cm)	Breakup category
Reported bridge damage										
1925	12 Feb	36,000	1020	6	100	625	20	18.3	0.20	3
1929	24 Mar	31,100	881	11	4600	445	48	28.1	0.69	1
1936	13 Mar	45,100	1280	1	600	790	3	20.0	4.80	3
1938	25 Mar	34,800	985	8	1900	585	26	56.8	0.05	3
1977	14 Mar	43,100	1220	2	900	741	7	59.4	3.89	3
No reported bridge damage										
1927	20 Mar	34,000	963	9	3900	580	29	59.7	0.53	1
1945	22 Mar	40,200	1140	3	4600	712	12	50.6	2.62	1
1946	9 Mar	31,000	878	12	800	744	6	34.4	2.92	2
1964	6 Mar	35,000	991	7	400	618	23	24.7	4.14	2
1968	22 Mar	34,000	963	9	2100	736	8	36.7	3.73	1
1979	7 Mar	40,000	1130	4	1000	671	18	37.5	6.48	2
1981	21 Feb	38,400	1090	5	4500	565	31	43.9	1.52	1
1986	27 Jan	19,700	558	28	100	641	19	0.0	6.99	2

produce the highest breakup water levels provides the foundation for the development of ice control methods.

ANALYSIS OF HISTORICAL DATA

The date of a dynamic ice breakup can be estimated from the daily flow records of a gaging station. Following a winter period of consistent low flows, the discharge increases rapidly at breakup and attains a high peak value. We obtained the peak daily average discharge and date of dynamic ice breakup, if it occurred, from 65 years of records of the Connecticut River and the White River at the gaging stations. The Connecticut River discharges were ranked and the 12 highest breakup discharge years, including all years of reported bridge damage, are given in Table 1. Several high discharge events occurred without significant rainfall at nearby Hanover, New Hampshire, indicating nonuniform rainfall on the drainage basin and a potentially important snowmelt contribution to the flow.

Accumulated freezing degree days is a measure that allows us to compare the ice production potential of a river between years. We calculated the total number of freezing degree days at Hanover, for the period of November through February, for 62 winters beginning in 1924–25. These values and the relative "coldness" ranking are given in Table 1 for each high breakup discharge year and for 1985–86. Of this group only 1936, 1946, 1968 and 1977 rank among the more severely cold winters, indicating that Connecticut River ice production in even an average winter is sufficient to pose a threat to the Cornish-Windsor bridge at breakup.

A prior midwinter breakup of the White River with subsequent refreezing would maximize the ice supply to the reach from the primary external source. Combined with minimal midwinter ice movement on the Connecticut River, this larger ice supply increases the flood potential of the reach in the spring. Of the 12 large event years, the White River ice released on 19 and 2 January in 1929 and 1979, respectively. Also, possible or partial breakups occurred on 26 January 1938 and 7 January 1946. We conclude that this uncontrolled process contributes in some years to high water levels at spring breakup.

High air temperatures in the period immediately prior to breakup could cause significant melting of the ice cover, minimizing the flooding from a dynamic breakup. However, the pre-breakup temperature data presented in Table 1 do not indicate any correspondence between melting degree-days and high water levels at Windsor. This lack of correspondence indicates a minor contribution of air temperature to ice cover decay, concurring with the model studies of Greene and Outcalt (1985). They found that water temperature was the most important thermodynamic variable controlling ice thickness, and that ice cover growth and decay were more sensitive to water temperature and flow velocity than to meteorological effects.

The process that we term "hydrothermal melting" refers to the melting of an ice cover from the heat contained in the flowing water, and the progressive failure of the weakened cover. Hydrothermal melting occurs when the net heat flux is into the river, typically corresponding to above-freezing daily average air tempera-

tures. Under these conditions the input of heat to the river is proportional to the area of open water (Marsh and Prowse 1987). The flow energy gradient, and consequently the hydrodynamic forces exerted on the ice, increase with discharge and cause progressive failure of the weakened cover. The additional open water area that results increases the available heat supply to the river and minimizes the time necessary for melting the ice. The discharge-days of hydrothermal melting for the period of rising discharge were computed from the daily average flow data and are given in Table 1. These data generally confirm that minimum hydrothermal melting of the ice and maximum water levels are compatible. An intact ice cover with the greatest resistance to breakup exists when the discharge increases rapidly to its peak, minimizing hydrothermal melt.

Because the Connecticut River is controlled, large flow changes occur daily on a time scale of hours. During the breakup period the river flow can increase suddenly, and in large events the river becomes uncontrolled. Because the flow conditions can change rapidly, daily average discharges do not provide the necessary information to adequately compare ice breakup in different years. For example, the instantaneous flow peak at the gage on 27 January 1986 of greater than 926 m^3/s is at least 66% greater than the daily average discharge. Further understanding of ice dynamics from the historical record requires hourly or more frequent resolution of the data.

In this report we will summarize the detailed historical White River and Connecticut River gage records. A more complete chronology is included in Appendix A. During ice breakup the rate of stage increase at each gage has frequently approached 2.0 m/hr. These high rates are associated with ice jam formation immediately downstream of the gage, or with ice breakup and jam release upstream. Stage recession rates of up to 1.4 m/hr have occurred in response to the release of an ice jam downstream. Discharge has increased at breakup by a factor of 20 or more during a 24-hr period on both rivers. An abrupt rise of the White River at the gage to a peak discharge greater than 300 m^3/s, sustained for several hours, indicates a complete breakup of the ice cover downstream to the Connecticut River confluence at an average speed of more than 1 m/s. The largest recorded daily average Connecticut River discharge during the normal period of ice breakup was 3260 m^3/s on 19 March 1936, a much larger event than any of the events to date that actually caused breakup.

The events listed in Table 1 fall into three general categories of ice breakup behavior. The first group of events (1927, 1929, 1945, 1968, 1981) exhibited high discharge with only gradual variations, and concurrent ice movement over a period of several days. A gradual and simultaneous breakup at several locations characterizes reduced energy gradient breakup behavior. Hydrothermal melting is effective under these conditions and rapidly reduces the quantity of ice participating in the breakup. The product of flow and duration exceeded 2000 m^3/s-days for each event in category 1, indicating significant potential for hydrothermal melting. The breakup was in an advanced stage when the peak discharge occurred, and water levels were generally moderate. The high water levels associated with bridge damage do not readily follow from the available data for 1929. However, the probable White River ice breakup in January 1929 distinguishes this event from the others in the group, and together with a lack of continuous flow data could explain the apparent contradiction.

The events in the second group (1946, 1964, 1979) each included the formation of a persistent upstream ice jam. The eventual release of the White River ice jam in 1964 produced the highest water levels since at least the 1920s at White River Junction, Vermont (pers. comm. with R.W. Lehman, Town Manager, Hartford, Vermont, 1986) and two spans of a three-span bridge over the White River were destroyed. This short-duration, extremely high flow input was not supplemented by a rising Connecticut River and experienced significant attenuation prior to arriving at Windsor. In 1946 and 1979 ice jams near the Connecticut River gaging station persisted for about 35 and 48 hr, respectively. The delay of the ice from the White River and the upstream reach of the Connecticut River provided an opportunity for breakup downstream to proceed with a smaller ice volume, effectively increasing the channel capacity. Therefore, the breakup at Windsor developed lower water levels than would have occurred if a single breaking front had advanced rapidly downstream.

The third group of events (1925, 1936, 1938, 1977) includes most years of reported bridge damage and the highest water levels at Windsor. In each case an abrupt White River rise deposited large quantities of ice in the Connecticut River. The intact and competent ice on the Connecticut River then began to fail as the discharge continued to increase rapidly, and the breakup traveled downstream. The largest quantities of ice together with a high peak discharge produce the highest river levels at breakup.

FIELD OBSERVATIONS

We regularly observed the ice conditions on the Connecticut River study reach throughout the winter of 1985–86. During December and January a generating unit at Wilder Dam was out of service, limiting releases to a maximum of 142 m^3/s. An ice jam commonly forms

over the winter in the reach near station 2. However, restricted peak flows limited the energy gradients and produced a uniform ice sheet. During normal two-unit flow releases, the reach upstream of station 1 is predominantly free of stable ice. The reduced unsteadiness and peak discharge during this period allowed the development of an ice cover of maximum extent upstream. An ice cover that has developed sufficient thickness can be very stable, as evidenced by the high discharges prior to breakup given in Table 1. After the breakup on 27 January we found large grounded ice floes exceeding 0.20 m thick at station 1 that were remnants of the upstream ice cover. Ice of this thickness is resistant to breakup at normal flow conditions. The presence of stable ice in this reach reduces the open water area below the dam, the heat loss from the river, and consequently, the total ice production.

The ice breakup on 27 January cleared the reach of ice from Wilder Dam to a point 2 km downstream of station 2. Following the breakup, a second ice sheet with uniform initial thickness formed near station 2. However, by the end of February, a massive jam had developed at this location; surface relief of 1.5 m indicated that the accumulation rested on the river bed at several places. The available freezing-degree-days were nearly equal during the development of the January ice sheet and February ice jam. However, the ice jam experienced much higher energy gradients with two-unit Wilder flow releases and a backwater effect from Bellows Falls Dam that was not present during the formation of the uniform sheet.

The February ice jam at station 2 blocked much of the channel and restricted the flow. The primary cause of the jam formation, evident from the ground and from low-altitude aerial photographs, was shoving and piling of thin ice. Mechanisms that contributed to the ice jam development were the increased ice supply from daily formation and breakup of thin ice upstream of station 1, and maximum open water area causing increased ice production. Ice deposition in the jam was initially favored by its location at the head of the backwater, and later by the channel blockage that developed.

Table 2. Controlled flow release test schedules at Wilder Dam, mean discharge of the White River during the monitoring period, and base flow of the Connecticut River at Bellows Falls Dam. **A release pattern number identifies each test schedule.**

	Release pattern number				
	1	*1*	*2*	*3*	*3*
	Discharge (m^3/s)				
Time	*15 Oct*	*26 Feb*	*27 Feb*	*28 Feb*	*16 Oct*
0000–0700	21	21	21	21	22
0700–0800	85	85	142	21	283
0800–0900	170	170	142	283	283
0900–1000	255	255	21	283	283
1000–1100	255	255	21	283	283
1100–1200	255	255	142	283	283
1200–1300	21	21	142	283	22
1300–1700	21	21	21	21	22
1700–2400	open	open	open	open	open
Mean discharge					
White River	27	22	19	19	43
Connecticut River at Bellows Falls Dam	75	59	55	54	134

Controlled release tests

The purpose of the controlled release tests on the Connecticut River was to obtain a comprehensive data set describing river wave behavior. Important physical parameters include the low flow hydraulic gradient and wave celerity between measurement stations, and the rate of stage increase, the wave front steepness, and the wave amplitude at several stations. In this discussion, wave celerity refers to the translation speed in the downstream direction. These data are then used in several ways: 1) to compare flow releases with different peak discharges and rates of increase to the peak discharge; 2) to determine the effects of the ice cover on river response, 3) to compare the response of different reaches and locations, especially noting changes with distance downstream; and 4) to compare the response of the Connecticut River with other rivers having different characteristics. This analysis provides us with an improved understanding of the parameters affecting the response of the river. These data also enable the calibration and verification of numerical models, providing additional testing of ice breakup theory with the goal of controlling ice breakup.

We conducted these tests on the Connecticut River on 15 October 1985 and 16 October 1987 during open water conditions, and on 26–28 February 1986 when ice was present in the river. A flow release schedule for each test day was specified for Wilder Dam (Table 2), while the discharge at Bellows Falls Dam was varied as required to maintain a constant headwater elevation. The schedules vary the rate of rise and the peak flow, and examine river response to successive releases spaced 2 hr apart. The mean discharge of the White River and the base flow of the Connecticut River at Bellows Falls Dam given in Table 2 are indicative of the relative inflows to the reach from local drainages. The open water tests had significantly higher inflows than the winter tests.

The ice regime of the river in our study reach was not changed substantially over the course of the winter test (Fig. 2). The few areas of solid shorefast ice that existed in sheltered bends of the river upstream of station 1 re-

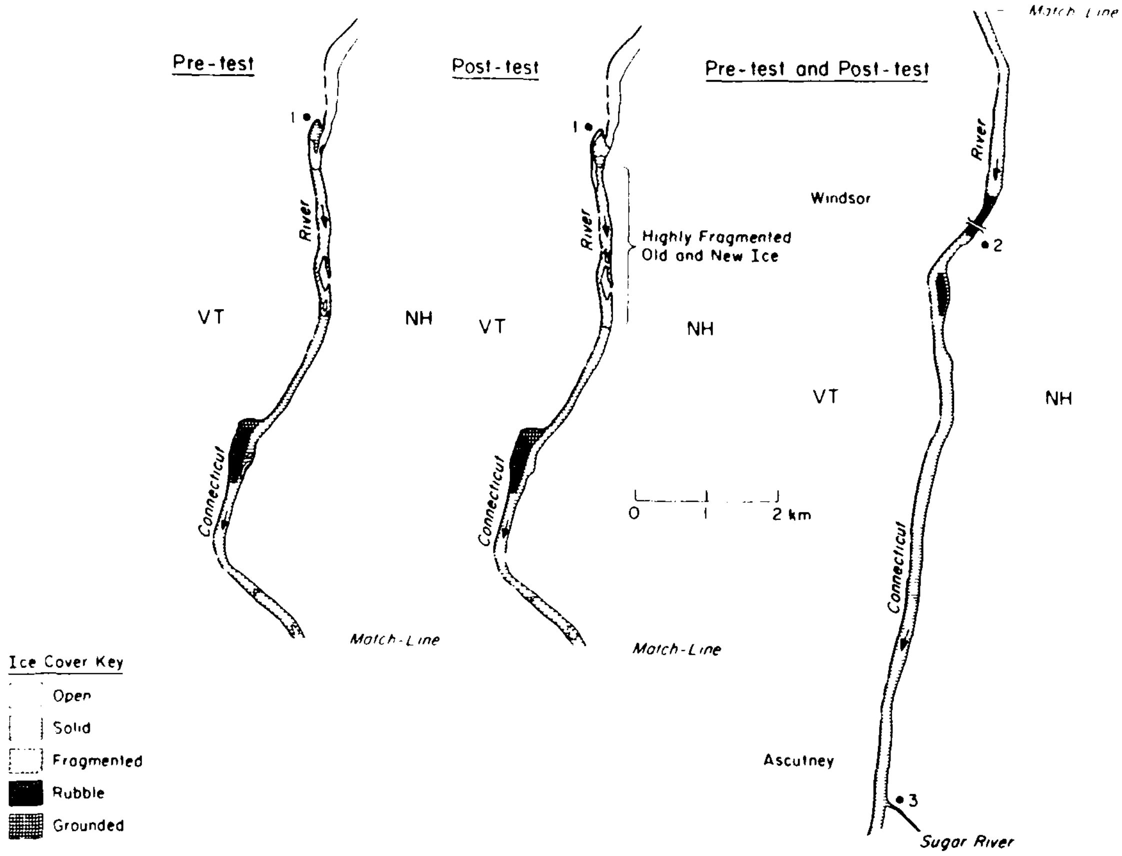

Figure 2. Ice conditions in the reach between stations 1 and 3, both prior to and following the winter test. The ice cover between stations 2 and 3 remained unchanged over the period of the test.

mained. Low overnight temperatures prior to each test day of –12°, –17° and –22°C, respectively, caused the growth of large expanses of thin ice in this largely open water reach. This ice was repeatedly broken, transported and deposited downstream of station 1 by the scheduled Wilder Dam release. The ice conditions in the 2-km reach downstream of station 1 became more highly fragmented over the duration of the test and the quantity of ice increased as a result of the supply from upstream. The surface appearance of the ice farther downstream was unchanged over the test. Accumulations of broken sheet ice near station 2 were several meters thick. The sheet ice thickness in sheltered areas near the right bank at station 2 was 0.56 m, and at station 3 the ice was solid and relatively uniform with a thickness of 0.53 m.

Stage hydrographs relative to mean sea level (msl) measured during the fall and winter tests are presented in Figure 3. The headwater elevation at the Bellows Falls Dam averaged 88.50 m above msl for the open water test on 15 October, and 88.27 m above msl for all remaining tests. A much larger stage response occurred at all stations with two-unit, compared to single-unit, Wilder Dam releases, and the disparity increased with distance downstream. Also, the rates of stage increase with time downstream of station 1 followed this same trend. These rates reflect the water surface and energy gradients, and indicate a direct relationship between the hydrodynamic forces on the ice cover and the magnitude oı the increase in discharge. These results are consistent with our observations at station 2 of single-unit releases and ice sheet development, or two-unit releases and ice jam development. The general increase in stage is evident at the downstream stations during ice cover conditions. Although the mean winter Bellows Falls headwater was 0.23 m lower than that of the initial open water test, the ice caused low flow stage increases of 1.4 m and 2.0 m at stations 2 and 3, respectively. The common water surface drawdowns at stations 2 and 3 early in each winter test day and the small stage difference between these stations indicate continuous backwater and a significant upstream extension of the Bellows Falls pool as a result of the ice cover.

The hydraulic gradients at low discharge presented in Table 3 were obtained for the reaches between measurement locations. The change in river stage of 2.2 m across the Sumner Falls rapids is nearly constant over the normal range of open water flow conditions. This abrupt local fall was not included in the calculation of the hydraulic gradients. The kilometer positions of the measurement locations listed in the table are based on a local system with distance increasing upstream from

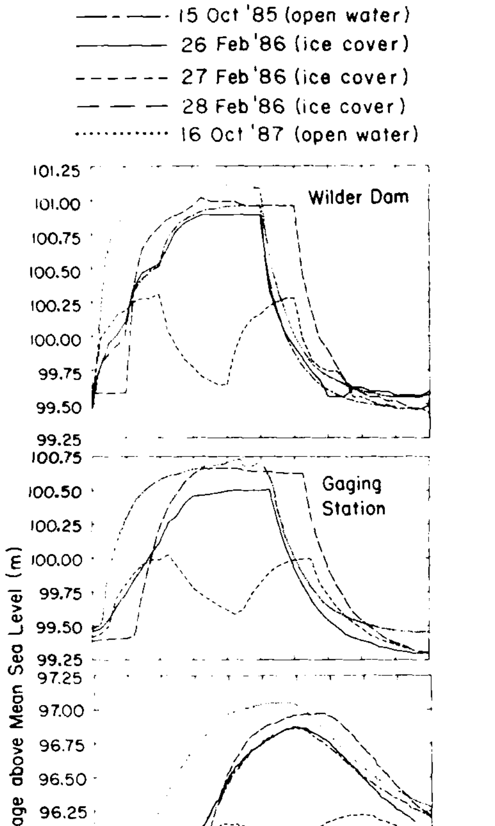

Figure 3. Measured stage during the fall and winter tests as a function of time at several locations in the study reach. The stage data at stations 2 and 3 required separate vertical scales to present both the ice cover and the open water results.

Bellows Falls Dam. Upstream of station 1 the winter gradient was somewhat larger than the open water gradient because of the spotty presence of ice. The river downstream of station 3 was completely ice covered and that gradient was significantly larger than it was during open water. This increased winter gradient downstream caused the head of the Bellows Falls backwater to shift from an open water location at about 2 km downstream of station 2 to an ice-affected location approximately 3 km upstream of station 2. As a result, the overall hydraulic gradient between stations 2 and 3 was reduced. Conversely, a 0.23-m reduction of the Bellows Falls headwater between the open water tests shifted the head of the backwater downstream. The open water data (Fig. 3) at station 3 reveal that the lower headwater elevation increased both the wave amplitude and the rate of stage increase, corresponding to an increased energy gradient of 0.20×10^{-3} between stations 2 and 3. The backwater at station 2 reduced the winter gradient between stations 1 and 2 relative to that during ice-free conditions.

Wave celerity is obtained by timing the initial stage increase between measurement locations. The relatively small quantity of ice upstream of station 1 on the morning of each test day did not affect wave celerity (Table 3). Higher open water celerities relative to those in winter were caused by higher local inflows and higher initial releases. Farther downstream, where a complete and stable ice cover existed, wave celerity was reduced substantially relative to open water conditions. Consistent wave celerity reduction in the presence of ice was also measured on the Hudson River by Ferrick et al. (1986a). Celerities in flat, pooled reaches that were significantly greater than those in sloped and freely flowing reaches, and celerity increases with increasing peak discharge were common to both rivers.

The rate of stage increase at a given location varies continuously during the passage of a river wave. Table 4 presents maximum rates of stage increase for 15-minute periods at the measurement locations on the Connecticut River. Upstream of station 1 the higher base flow present during the open water tests caused a greater reduction with distance of the maximum rate of stage increase than in winter. However, the permanent ice cover downstream of station 1 reversed this comparison. The significant reduction of the maximum rate of stage increase due to the ice cover was also reported by Ferrick et al. (1986a) for the Hudson River. Higher rates of stage increase with higher peak discharge, and a reduction in these rates with downstream distance occurred in both rivers.

The wave front steepness estimates presented in Table 5 were obtained by dividing the rate of stage increase by the wave celerity. Estimated wave steepness

Table 3. Wave celerity variations with release discharge and cover condition, and comparison of low discharge hydraulic gradients by reach and cover condition.

		Low discharge hydraulic gradient ($\times 10^{3}$)		*Celerity (m/s)*						
				1	*1*		*2*	*3*	*3*	
Location	*km*	*open*	*ice*	*open*	*ice*	*ratio*	*ice*	*open*	*ice*	*ratio*
Wilder	68.2									
		0.014	0.056	2.23	2.23	1.00	2.02	2.82	2.49	1.13
Gaging sta.	65.6									
		0.30	0.32	2.13	1.99	1.07	2.04	2.94	2.55	1.15
Sta. 1	54.1									
		0.31	0.25	1.43	1.01	1.41	0.88	1.87	1.18	1.59
Sta. 2	42.6									
		0.15	0.093	2.03	1.35	1.51	1.19	2.26	1.52	1.48
Sta. 3	34.2									
		0.004	0.070	3.24	—	—	—	3.26	2.26	1.44
Bellows Falls	0.0									

Table 4. Variations in the maximum rate of stage increase with release discharge and cover condition.

		Max. rate of stage increase (m/min)						
		1	*1*		*2*	*3*	*3*	
Location	*km*	*open*	*ice*	*ratio*	*ice*	*open*	*ice*	*ratio*
Wilder	68.2	0.023	0.020	1.13	0.031	0.059	0.053	1.11
Gaging sta.	65.6	—	0.0091	—	0.011	0.026	0.024	1.11
Sta. 1	54.1	0.0069	0.0079	0.88	0.0053	0.0087	0.012	0.74
Sta. 2	42.6	0.0060	0.0062	0.97	0.00087	0.0065	0.0068	0.96
Sta. 3	34.2	0.0034	0.0034	1.00	0.00000	0.0046	0.0040	1.14

Table 5. Variations in the estimated wave front steepness with release discharge and cover condition.

		Estimated wave front steepness ($m/m \times 10^{3}$)						
		1	*1*		*2*	*3*	*3*	
Location	*km*	*open*	*ice*	*ratio*	*ice*	*open*	*ice*	*ratio*
Wilder	68.2	0.17	0.15	1.13	0.26	0.35	0.35	0.98
Gaging sta.	65.6	—	0.068	—	0.091	0.15	0.16	0.96
Sta. 1	54.1	0.054	0.066	0.82	0.043	0.049	0.077	0.64
Sta. 2	42.6	0.070	0.10	0.68	0.016	0.058	0.096	0.60
Sta. 3	34.2	0.028	0.042	0.66	0.000	0.034	0.044	0.77

Table 6. Wave amplitude variations with release discharge and cover condition.

		Wave amplitude (m)						
		1	*1*		*2*	*3*	*3*	
Location	*km*	*open*	*ice*	*ratio*	*ice*	*open*	*ice*	*ratio*
Wilder	68.2	1.50	1.29	1.16	0.79	1.58	1.40	1.13
Gaging sta.	65.6	—	1.05	—	0.61	1.21	1.26	0.96
Sta. 1	54.1	1.01	1.11	0.91	0.49	1.03	1.19	0.87
Sta. 2	42.6	0.75	0.88	0.85	0.17*	0.86	1.01	0.85
Sta. 3	34.2	0.51	0.51*	—	0.00	0.79	0.60*	—

* Stage rising beyond measurement period.

increased with peak discharge and generally decreased with distance downstream. The presence of an ice cover caused a larger increase in estimated wave steepness between stations 1 and 2 than occurred with open water. However, between stations 2 and 3 wave steepness reductions were either comparable or were larger with ice present. Ferrick et al. (1986a) reported larger wave steepness reductions in the Hudson River with ice present. Possible explanations for the different effects of the ice on wave steepness include a difference in wave type (Ferrick 1987), and imprecision of this steepness estimate.

Wave amplitude is the stage increase at a given location obtained from measurements immediately prior to wave arrival and at the wave peak. Wave amplitude and amplitude attenuation are a reflection of the channel capacity, a function of the geometric and hydraulic characteristics of the river. Wave amplitude (Table 6) increases with peak discharge and decreases with distance downstream as a result of attenuation. With ice in the river the flow resistance is increased, causing higher wave amplitudes relative to identical open water flow conditions.

January 1986 ice breakup

The air temperatures in the week prior to the ice breakup that occurred on 27 January were seasonally low, totaling 61 freezing degree-days. Because of the low temperatures, the ice in the reach was strong, with thicknesses ranging between 0.3 and 0.5 m. A rainfall on 26–27 January of more than 6 cm provided the source of the inflow that eventually led to the breakup.

River stage and discharge data at Wilder Dam, and at the Connecticut River and White River gaging stations, are presented in Figures 4 and 5, respectively, for 27 January. The flow release at Wilder increased throughout the day, attaining a peak of more than 10 times the initial flow. The elevation of the headwater at Bellows Falls Dam was held constant at 87.97 m above msl, the minimum pool of the normal operation range. The stage-discharge relationship at a gaging station is developed during relatively steady flow conditions. However, the flow is rapidly varied during a dynamic breakup and the rating curve can significantly underpredict the discharge. The unsteady flow correction for the White River gage is generally minor because of the large stream bed gradient. The Connecticut River bed gradient is much smaller and unsteady flow corrections can be important. In addition, the Connecticut River stage was ice affected early on 27 January, requiring an additional correction. Therefore, the Connecticut River gage discharge record is only an estimate of actual instantaneous discharge. This conclusion is supported by continuity considerations between the White River inflow, Wilder Dam release and the Connecticut River gage discharge. The 310-m^3/s flow increase at WRK 11.9 was sustained for 2 hr. At the confluence the peak inflow should have been even higher and/or the duration longer with the additional water release from storage during the complete breakup of the lower White River. In contrast, the gage showed a peak flow increase of only 220 m^3/s above the Wilder discharge, which was sustained for less than 1 hr before subsiding.

Knowledge of the energy gradient at breakup is essential for ice management. However, sufficient data to characterize the breakup are not available, and the energy gradient at station 2 will be estimated on the basis of limited on-site data and known upstream flow conditions. Steady-flow open water data yield a Manning's roughness coefficient of 0.025 for the reach between stations 1 and 2. Variable ice roughness and thickness, and sparsity of data introduce uncertainty into a combined ice-bed roughness determination. With an estimated mean ice thickness we obtained a combined roughness for the February test that is comparable to the open water value. At 1400 hr, prior to breakup at station 2, the estimated discharge, river width and mean depth beneath the ice were 740 m^3/s, 150 m and 4.0 m, respectively, yielding a mean flow velocity of 1.23 m/s and a dynamic wave celerity of 7.5 m/s. The corresponding energy gradient estimate is 0.38×10^{-3} but a combined roughness of 0.023 reduces this estimate to 0.31×10^{-3}, the streambed gradient. The increased flow resistance from the static ice cover increased the flow depth by 1 m relative to open water, effectively holding this volume in storage.

The stage decrease between 1300 and 1400 hr at the gaging station with steady flow conditions at Wilder and from the White River indicated an ice release downstream of the gage that was timed properly to have initiated the downstream breakup. The breakup began upstream and progressed rapidly downstream as a single feature with no ice source external to the reach itself. With wave arrival at station 2 the entire ice sheet was forced downstream. Behind the breaking front, which we define as the location of initial ice motion, the ice sheet was transformed into a mosaic of plates with initial size and shape determined by the pattern of preexisting fractures. The breaking front moved ahead of the ice plates that were not directly involved in the breakup. Plate size was reduced with distance as a result of impacts with the banks, with obstructions in the river (i.e. bridge piers) and with each other. The large plate collisions immediately behind the breaking front restricted their movement, while farther upstream smaller brash moved freely. The difference between these speeds resulted in a zone of ice convergence and a rapid transition from plates to ice rubble. This second ice

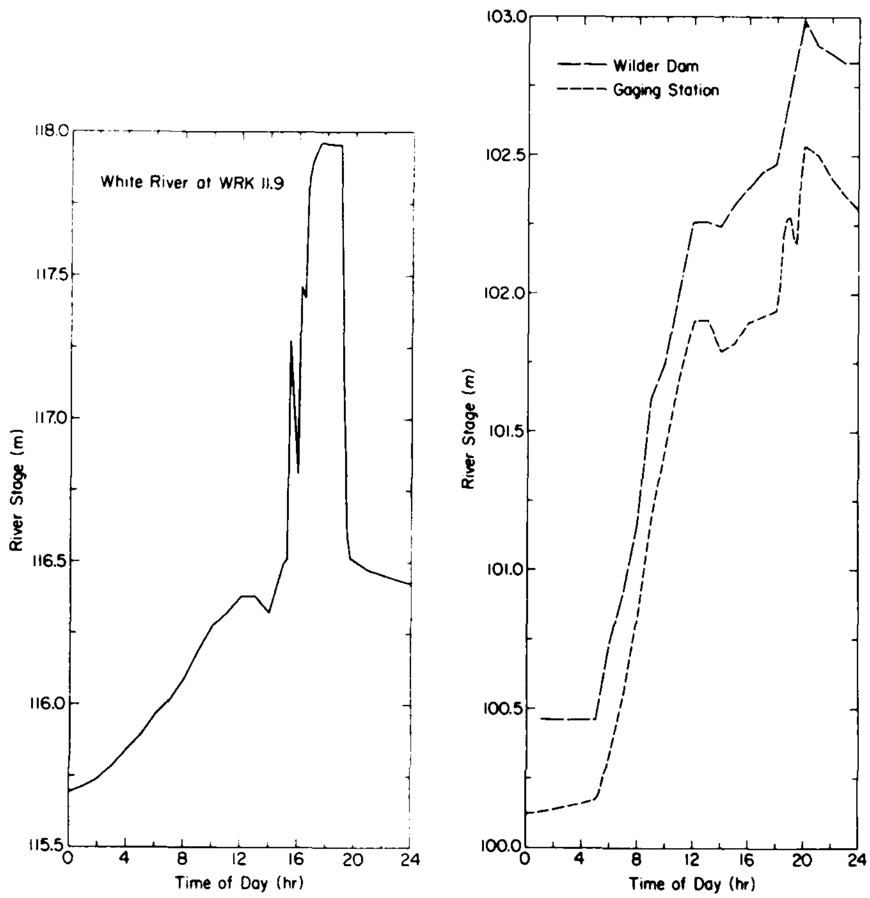

Figure 4. River stage variations with time for 27 January 1986 at the White River gage WRK 11.9, the Connecticut River gage and the tailwater at Wilder Dam.

rubble front followed the breaking front at station 2 by a distance of about 2 km. As the breakup progressed into the Bellows Falls backwater, the energy gradient diminished and became insufficient to sustain the ice breakup at about 40.5 km. The fragmented ice visible from station 2 was motionless at 1530 hr, coincident with the peak stage. A rapid stage subsidence of 0.3 m by 1600 hr indicated the short-duration characteristic of the river wave resulting from the breakup of the reach. The stage peak at station 2 attained in this event was 95.11 m above msl, corresponding to 3.71 m of freeboard at the Cornish–Windsor bridge.

Our observations and the ice breakup theory of Ferrick et al. (1986b) indicate that the speed of the breaking front cannot be greater than the dynamic wave celerity. The flow resistance decreases rapidly as the ice begins to move. Ice moving at the flow velocity causes negligible resistance to the flow. If the acceleration of the ice plates is instantaneous, the stored water is released at the speed of the breaking front. This maximum contribution to the river wave yields an upper bound on the energy gradient. The breaking front speed, estimated at 5 m/s near station 2, and the peak gradients that develop are related. The rapid movement of the front indicates that large energy gradient increases were not necessary to initiate breakup (Ferrick et al. 1986b). Therefore, our estimate of the energy gradient at breakup is in the range of 0.5 to 0.7×10^{-3}. Both the energy gradient and the mean ice thickness are an order of magnitude greater than those of the midwinter Hudson River breakup (Ferrick et al. 1986a). However, the physical mechanisms characterizing breakup in these two cases were different.

The second stage of the breakup event was caused by

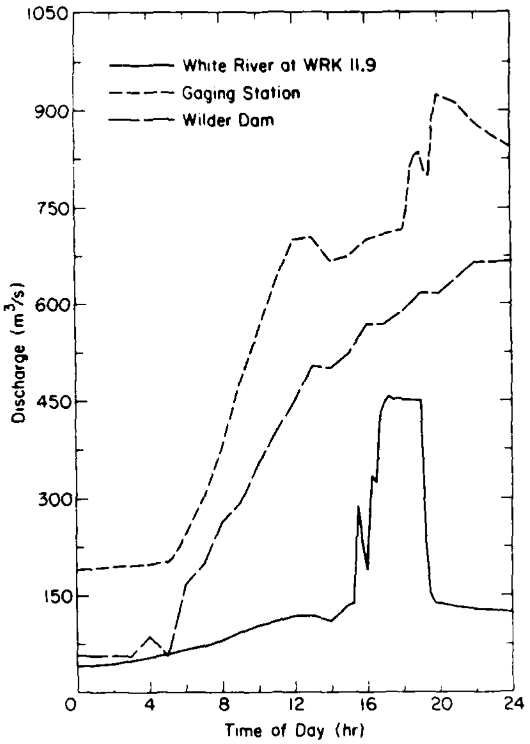

Figure 5. Riverflow variations with time for 27 January 1986 for the tailwater at Wilder Dam and estimates based on the rating tables for the gages on the White and Connecticut Rivers.

the river wave from the breakup of the White River. At 1615 hr a large wave from ice breakup upstream passed the gaging station at WRK 11.9. Observations of the ice sheet in the lower 3 km of the White River revealed intact ice at 1700 hr. The flow in the Connecticut River at the gaging station began increasing at 1800 hr in response to the rising White River. The White River ice run accompanied by sharply increasing flow arrived at the gage at 1945 hr, followed closely by the peak discharge at 2000 hr. The speed of the breakup in the lower White River averaged 1 m/s, requiring just over 3 hr to progress the 11.9 km from the gage to the Connecticut River confluence. The speed of the wave front was constrained by the velocity of the water, which in turn was limited by the presence of the ice. The relatively steep bed slope (0.0013), the modest initial flow with an abrupt transition to a high peak flow, and the high roughness of the ice-covered channel caused the relatively low wave celerity.

With the arrival at Windsor of the breakup wave from the White River, the energy gradient increased, causing the ice jam to fail and travel about 16 km farther downstream into the Bellows Falls backwater. Wave movement and corresponding changes in energy gradient are much faster than the velocity of ice floes in relatively deep open water. Moving ice downstream of Windsor was observed several hours before the ice from the White River could have arrived. The subsequent ice jam, again caused by an insufficient energy gradient, was about 16 km long, including the White River ice deposited at the head of the jam upstream. The post-breakup arrival of the White River ice in this event was similar to the historical events that were characterized by persistent upstream ice jams.

CONNECTICUT RIVER ICE CONTROL

The highest water levels on the Connecticut River near Windsor, Vermont, occur with the following critical sequence of ice breakup events. The White River rises abruptly to a high peak, depositing large quantities of ice in the Connecticut River. Meanwhile, the ice on the Connecticut River is competent and intact, and the combined discharge continues to rapidly increase toward a peak daily average flow in excess of 1200 m³/s. The entire volume of ice is involved when the Connecticut River ice run occurs, and the channel blockage that develops combined with the high discharge causes extreme water levels.

The predictable timing of a dynamic ice breakup works to the advantage of any ice control method. The data in Table 1 indicate that 10 of the 12 largest breakup events occurred in a 2½-week period in March. All of these events closely follow and occur in response to a significant rainfall. With river regulation the available methods of ice control are to 1) minimize ice production over the winter, 2) reduce the quantity of ice prior to breakup, and 3) disrupt the natural White River–Connecticut River breakup sequence, reduce the peak breakup discharge, and minimize the river stage during breakup.

Minimizing ice production

The total ice production of a river can be minimized by river regulation during the period of ice growth. The basic concept of ice control is that by enabling a uniform and stable ice cover to develop with the maximum possible areal extent, the water is insulated from midwinter air temperatures. The test and observational data indicate that a low (< 85 m³/s) steady discharge should be sustained until the maximum ice cover extent develops with thicknesses generally greater than 20 cm. If fluctuations of the Wilder Dam release are necessary, the rate of flow increase will limit the energy gradient increase downstream. Tables 3 and 5 indicate that flow increases in increments of less than 60 m³/s per hour

constrain the energy gradient increase to less than 15% above the low discharge gradient downstream of the gaging station. A cold ice cover that is several centimeters thick will likely resist limited fluctuations at releases lower than a single unit maximum. Rapid decreases in the flow release do not increase the energy gradient and need not be regulated. The 20-cm thickness requirement ensures that the cover will not readily break up when normal operations resume. This method avoids ice jamming during freezeup that is caused by repeated shoving and breakup of the immature ice cover, and minimizes the *total* production of ice by minimizing the surface area of open water contributing to the heat loss. The process is repeated if a midwinter ice breakup occurs.

Hydrothermal melting

The concept of managing the heat balance of the river in early spring is identical to that in midwinter, but the control operations are reversed. In spring the air temperature and solar insolation increase and the net heat flux is into the river. The ice cover favorably minimizes heat loss in winter, but unfavorably minimizes heat gain in early spring. A method of ice control in the early spring is to maximize the ice melt prior to a dynamic breakup event. The melt and gradual breakup of ice in the upstream part of the reach when heat flux is into the river increases the area of open water, the heat gain by the river, and the rate of ice melt. The historical data suggest that high water levels are avoided when sustained moderate discharge and moderating above-freezing air temperatures produce significant hydrothermal melting and weakening of the ice cover. Marsh and Prowse (1987) observed a very high heat flux from the water to the ice and a rapid water temperature decay in a few kilometers from open water into an ice accumulation. These results suggest that heat transfer from the water occurs rapidly after encountering stationary ice, concentrating the ice melt near the leading edge. The subsequent collapse of the thinned ice cover occurs at a reduced energy gradient.

The ice sheet in the Connecticut River upstream of station 2 is about 15 km in length, 125 m wide and 0.5 m thick. If we assume adiabatic melting of the ice by heat contained in the water, the change in enthalpy or internal energy of the water must balance the latent heat of fusion of the ice mass. The product of water volume and temperature needed to melt all the ice in the reach upstream is 6.9×10^7 m^3-°C. With a constant water temperature of 0.2 °C at the upstream edge of the ice sheet, a release of 280 m^3/s for 14 days would be required. However, water temperature increases reduce the necessary volume of water proportionally. The full 14-day release yields 3900 m^3/s-days of hydrothermal melting (Table 1). Historical breakups with a comparable melt characteristic (category 1) were not generally associated with high water levels.

We recommend a sustained continuous discharge of at least 280 m^3/s from Wilder Dam for between 7 and 10 days with minimum Bellows Falls headwater elevation, coordinated with daily average air temperatures of 0°C or higher in late February and early March. If the available water is insufficient for a continuous release, hydrothermal melting can be optimized by coordinating releases with maximum solar insolation and air temperature. Sufficient lead time usually exists to greatly reduce the quantity of ice in the river at the time of breakup.

Controlled ice breakup

Controlled ice cover breakup is a method that disrupts the critical combination of events that causes extreme water levels and can be implemented on the relatively short notice of a 1- to 2-day weather forecast. The concept is to cause the Connecticut River to break up prior to the natural event at a lower stage and discharge, and in advance of the White River breakup. Another benefit is that the early release of water from upstream that causes the breakup also creates storage and reduces the eventual peak discharge. Both the observed breakup on 27 January 1986 and the analysis of the historical records clearly indicate the flood reduction advantage of separating the White River and Connecticut River breakups

Precise release patterns from Wilder Dam that will produce a controlled breakup have not yet been developed, but guidance is available from the conditions of the January 1986 event. Abrupt releases of several hours in duration provide maximum energy gradients and icebreaking capability with a minimum volume of water released and at minimum river stage. Lowering the Bellows Falls headwater from the winter test elevation of 88.27 m to 87.35 m above msl would shift the head of the pool downstream toward its open water location. The minimum possible Bellows Falls headwater elevation increases the attainable energy gradients downstream from station 2, limiting the size of the upstream release needed for effective ice breaking. This method of ice control for the Windsor area would not change the ice breakup behavior in the lower 30 km of the Bellows Falls pool. From that location upstream, the controlled breakup would occur earlier, and at a reduced stage and discharge, than the natural event. Releases designed to produce hydrothermal melting would be initiated immediately following this breakup. The combination of high roughness of the ice jam downstream and the large area of open water to maximize heat gain upstream yields the highest possible rate of melt of the

accumulated ice. Additional laboratory modeling to develop a test plan and a successful field trial of this plan are necessary before attempting a controlled ice breakup.

CONCLUSIONS

The potential exists for a much larger ice breakup event on the Connecticut River than has occurred in the historical record. Therefore, raising the Cornish–Windsor bridge elevation by 0.91 m (3 ft) will not ensure a safe structure during breakup without additional ice control measures. Three basic methods have been identified that use river regulation to minimize the probability of breakup flooding and damage or loss of the bridge. The use of all these methods, to the largest extent possible, would produce minimum river stage during the period; however, all have associated power production costs. Controlled Connecticut River ice breakup is the only alternative for ice management that provides an immediate response capability. The development of a well-founded test plan for a controlled ice breakup could be accomplished with available data using the theory of Ferrick et al. (1986b). A successful field trial of this plan is required before ice breakup control can be implemented on the Connecticut River.

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APPENDIX A: DETAILED ICE BREAKUP CHRONOLOGY

1925, 1929, 1936

There were no continuous recording gages on the Connecticut and White Rivers prior to 1930. The rapid rise of the river to a high peak discharge and minimal hydrothermal melting produced the flooding in 1925. In contrast, the high water levels associated with bridge damage do not readily follow from the available data for 1929. The continuous recording gage on the Connecticut River was not operating during the 1936 breakup period. Bihourly data from the White River gage indicate a large stage and discharge increase on 12 March 1936 from 59 m^3/s to a peak flow of 629 m^3/s, clearing the river of ice. The peak Connecticut River discharge occurred the next day, with White River ice present in the channel. The release of White River ice with intact ice on the Connecticut River and a rapidly rising discharge to a high peak is the combination of events that produces the highest water levels in the Windsor area. A second flood peaked on 19 March 1936 at the Connecticut River gage with a daily average discharge of 3260 m^3/s. Extreme flood damage and loss of the covered bridge would have been probable if this flood had encountered an intact ice cover.

1938

In 1938 the stage on both the White and Connecticut Rivers began rising on 21 March. The stage on the White River did not attain levels indicative of breakup until after a sharp rise on the evening of 23 March. The stage peak on 24 March was followed by a rapid recession corresponding to the probable clearance of ice from the river. A significant rise at the Connecticut River gage began late on 23 March, with ice motion evident on 24 and 25 March. The conclusions drawn from these data are that the period of rising discharge through levels with significant melt or breakup capability is much shorter than estimates from daily records would indicate, and again, the sequence of White River ice release immediately prior to Connecticut River ice movement produces the highest water levels.

1945

Rising stage and ice movement began at the gaging stations on both the White and Connecticut Rivers on 15 March 1945. The water levels at both locations continued to rise over the next several days, and the continuation of ice movement on the Connecticut River is evident in the records. A very high peak discharge occurred on the Connecticut on 22 March, but most of the ice was already past with high discharge sustained for several days prior, causing extensive hydrothermal melt and progressive breakup. The receding stage on 23 March was not affected by ice.

1946

A large-amplitude short-period river wave peaked at the White River gage on 7 March 1946 at 2130, causing the ice breakup in the lower White River. At 0200 on 8 March the steadily rising Connecticut River began a dramatic rise with a peak rate of stage increase of 1.2 m/hr as the ice and wave front arrived from the White River. In the next 2 to 3 hr an ice jam formed in the reach downstream of the gage that remained in place for about 35 hr until releasing suddenly on 9 March, indicated by a stage decrease at the gage of 1.4 m in 1 hr. The delay of the upstream ice in the jam allowed the breakup downstream to proceed with a smaller ice volume. The rapid jam release initiated a wave with a flow peak that apparently was insufficient to cause flooding and bridge damage at Windsor.

1964

An abrupt stage increase at the White River gage of 1.6 m in 1 hr began on 5 March 1964 at 1130 hr. It was associated with ice breakup, but did not have sufficient amplitude and duration to cause a breakup to propagate down to the river mouth. A second stage rise began at 1630 hr and reached a peak at the gage at 0200 hr on 6 March that was 1.1 m higher than the initial wave. This second wave caused the release of a major ice jam that had formed at 2.4 km from the mouth (pers. comm. with R.W. Lehman, Town Manager, Hartford, Vermont, 1986). The jam release arrived at the mouth at 2400 hr producing the highest water levels since at least the 1920s at White River Junction, Vermont, and two spans of a three-span bridge over the White River were destroyed. Stage at the Connecticut River gage indicated normal power production at Wilder Dam early on 5 March. A second rise above normal stages began at 1100 hr. Two minor ice shoves preceded a larger movement of ice downstream of the gage at 1830 hr. After that time ice movement is evident in the stage record. The release of the White River jam introduced a very high but short-duration wave into the Connecticut, peaking at 0100 on 6 March at the gage. Prior to the peak the stage at the gaging station increased 1.8 m in 1 hr. The arrival of the primary White River ice at the Connecticut River gage lagged the initial ice movement there by several hours. A stage recession of 2.7 m within 12 hr after the peak indicates rapidly diminishing flow,

ensuring significant wave peak attenuation downstream at Windsor.

1968

The White River gage records for March 1968 could not be located. The Connecticut River gage record indicates very low stages through noon on 17 March 1968. The river then began a gradual 4.5-m stage increase over the next 4.6 days, peaking at 0200 hr on 22 March. Fluctuating river stage, indicative of the presence or movement of ice, ends at 0600 on 22 March. The gradual flow increase that accompanied the 0.04-m/hr average stage increase probably caused progressive hydrothermal ice failure that did not produce damaging high water levels at Windsor.

1977

On 10 March 1977, the White River gage recorded a low typical winter stage until noon, followed by the beginning of a gradual rise. By the end of the day on 12 March the stage had risen 0.8 m and minor ice movements are indicated in the records. The Connecticut River gage was out of service in 1977. The flow release from Wilder Dam on 12 March varied between 0 and 266 m^3/s, all within normal turbine operating range. A significant jam formed downstream of the White River gage at 0230 hr on 13 March, persisting for about 12 hr. In response to rising discharge the jam released suddenly and was followed immediately by an abrupt peak that increased the stage 1.7 m in less than 1 hr. The peak at the gage, caused by ice breakup upstream, occurred at 1700 hr and was high enough to rapidly clear the lower river of ice. The Wilder Dam releases generally increased through the day on 13 March, initially exceeding turbine capacity at 1800 hr. Therefore, the White River ice arrived and probably jammed at intact and competent Connecticut River ice cover that evening. After the ice cleared from the White River, the discharge continued to increase, peaking at 0800 hr on 14 March at a flow of at least 656 m^3/s. The Wilder Dam releases also continued to increase on 14 March from 470 to 702 m^3/s. The combined discharge caused the breakup of the competent Connecticut River ice and produced extremely high water levels. During this event the Bellows Falls headwater elevation varied between 88.24 and 88.76 m above msl, higher than the normal minimum pool elevation of 87.97 m. The discharge records from the White and Connecticut Rivers indicate that the 1977 breakup was not the largest possible event. However, it also displays the combination of conditions that cause the highest water levels at breakup.

1979

An abrupt stage increase with a maximum rate of 1.3 m/hr began at the White River gage on 5 March 1979 and reached a pair of peaks just after midnight that correspond to ice movement. The Connecticut River gage indicated an abrupt 3.8-m rise in water elevation early on 6 March 1979, also with a maximum rate of increase of 1.3 m/hr. Part of the stage increase is from the formation of an ice jam downstream of the gage. At 0800 on 6 March the jam released and the river stage fell 1.2 m in 1 hr at the gage. Meanwhile, ice movement in the White River continued. A rise of 1.6 m in 1 hr peaked the White River at 1300 with a flow of at least 630 m^3/s. This wave would have totally cleared the lower White River of ice in 2–3 hr. The discharge in the White River remained moderately high after this wave passed until the morning of 7 March when it began a rapid recession. Unlike the 1977 event the high discharge from the White River was not sustained following its ice breakup. The Connecticut River stage remained high and ice affected on 7–8 March with a very high discharge, indicating extremely competent ice. This jam persisted for about 48 hr until 1800 hr on 8 March when the river stage at the gage began a significant decline that corresponded to the clearance of ice from its vicinity. Again, the lengthy delay of upstream ice in the jam probably reduced channel blockage and water levels during the breakup at Windsor

1981

Inspection of the hourly data for 1981 reveals a progressive breakup that occurred prior to 21 February and at lower discharges. Ice movement began on both rivers on 2 February. A stage increase on that date at the Connecticut River gage of 4.6 m with a 1.9 m/hr maximum rate of increase corresponds to the formation of a substantial ice jam. Ice jams were observed both above and below the gage on 4 February by the U.S. Geological Survey. Ice continued to produce a very high river stage through 11 February, but the discharge remained relatively low. The discharge then began rising in both rivers, and the White River peaked at the gaging station with a discharge exceeding 290 m^3/s at 0200 hr on 12 February. The ice remaining in the White River was deposited in the Connecticut River early that day. The persistent ice jam on the Connecticut released after 0400 hr, causing a 2.5-m stage decrease at the gaging station in 2 hours. The daily average flow in the Connecticut River for 12 February was 765 m^3/s, high enough when combined with a sudden ice jam release to extend the breakup well downstream of Windsor. Moderately high discharges continued until the peak on 21 February, ensuring negligible ice effects at that time.

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